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NAGW-1021

IN-46-CR

378

# Nonlinear Waves in the Winter Stratosphere of the Southern Hemisphere

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November 1988

(NASA-CR-186740) NONLINEAR WAVES IN THE WINTER STRATOSPHERE OF THE SOUTHERN HEMISPHERE (California Univ.) 37 p

N90-71250

Unclas

00/46 0292231

## Abstract

The variability of the winter circulation in the Southern Hemisphere at 10 mb is described in terms of the empirical orthogonal functions (EOFs) of the time series of geopotential height field anomalies covering the years 1979–1985. The patterns of the first three EOFs are similar except for a zonal shift. It is found that these EOFs are associated with large, sometimes eastward-traveling wavenumber one-type disturbances, which develop in some early winters (1980, 1984, 1985) over a preferred geographical location (south of Australia). It is also found that the fourth and fifth EOFs are associated with eastward moving zonal wavenumber two-type disturbances, which are generally active in late winter. No apparent relation is found between the variability of the stratospheric and tropospheric circulations of the Southern Hemisphere described by EOFs of geopotential height field anomalies at 10 mb and 500 mb.

An examination of the individual early winter episodes of large zonal wavenumber one-type disturbances reveals that the growth of the disturbances in the stratosphere is preceded, in some cases (1980 and 1985), by an amplification of the zonal wavenumber-one component of the flow in the troposphere. In other cases, there is no clear relationship between the evolution of the stratospheric and tropospheric flows. The stratospheric disturbances, therefore, can be generated by tropospheric forcing, but their evolution is dominated by local processes. Since the early-winter stratospheric disturbances are only found in years when the QBO is in its westerly phase, a possible relation with the Antarctic spring 'ozone hole' is considered.

## 1. Introduction

In the classical picture, the winter and spring circulations in the Southern Hemisphere (SH) stratosphere are relatively quiescent and highly disturbed, respectively, as expected from the predictions of the linear theory of Charney and Drazin (1961). According to this theory, the upward propagation of stationary planetary waves from the troposphere becomes more likely as the strong westerlies characteristic of the winter relax during spring. This picture of the stratospheric circulation, and its interpretation in terms of Charney-Drazin theory, has been challenged by the results of recent observational studies. Geller and Wu (1987) and Randel (1988) showed that, at high latitudes, the stationary and transient components of the stratospheric flow with zonal wavenumber one have amplitude maxima in early winter, as well as in spring.

The spring maximum in the component of the stratospheric flow with zonal wavenumber one represents the intensification of a quasi-stationary anticyclone (Mechoso et al., 1988). This anticyclone, which plays a crucial role in the breakdown of the cyclonic circumpolar vortex during the spring final warming, is one among several large-scale features of the SH stratospheric flow systematically observed during this season. In every October of the eight-year set analyzed by Mechoso et al., the monthly mean position of the anticyclone lies roughly between 90°E and 180°, directly above a pool of warm air in the lower stratosphere and a split of the westerly jet stream in the troposphere. Such a systematic character strongly suggests that these features are linked to surface boundary conditions. In addition, the numerical experiments on atmospheric sensitivity to orographic distributions performed by James (1988) are supportive of a link between the zonal asymmetry of the Antarctic orography and the split in the tropospheric westerly jet.

The focus of this paper is on the three-dimensional evolution of the circulation during early winter in the SH stratosphere. As in Mechoso et al. (1988) we use the

National Meteorological Center (NMC) stratospheric dataset to analyze the development of large-scale disturbances and discuss the generation mechanisms of these disturbances. To study its evolution, the flow on pressure surfaces is described in terms of two-dimensional structures, not just one-dimensional structures along latitude circles as in Fourier decomposition. These two-dimensional structures are the empirical orthogonal functions (EOFs) of the time series of geopotential height field anomalies.

EOF analysis has proven useful in several studies of low-frequency variability in the tropospheric circulation (Horel, 1981; Barnston and Livezey, 1987; Mo and Ghil, 1987). It has not yet been applied to research in stratospheric dynamics, since the datasets were too short to compile stable statistics. For the stratosphere, as for the troposphere, the first EOFs are expected to be dominated by large-scale patterns associated with well-defined dynamical features. It is reasonable at this point to attempt a preliminary statistical study of the stratospheric circulation since the NMC analyses for this region are now almost a decade long.

Given the questions raised recently (Geller and Wu, 1987; Randel, 1988) about the role of tropospheric forcing in the generation of the large-scale stratospheric disturbances, we follow a new approach in this study, and attempt to relate the evolution of the primary modes of variability, as revealed by the EOF analysis, in the stratosphere and the troposphere. Ghil (1987; 1988, Lectures V and VI) has recently explored systematically the connection between the apparently linear statistical tool of EOF analysis and nonlinear atmospheric dynamics. Other approaches to this problem used arguments of linear-wave theory and/or Fourier decomposition. Yamazaki and Mechoso (1985) inspected the time evolution of the Eliassen-Palm (E-P) flux in the SH during the spring final warming of 1979. The vertical component of the E-P flux measures the vertical propagation of wave activity in the framework of quasi-geostrophic linear-wave theory (Andrews and McIntyre, 1976; Boyd, 1976). They found that an event of enhanced planetary-wave activity observed in the stratosphere could be traced back in time and down to the troposphere. The same inspection performed for several other SH springs, however, shows that not all events of enhanced E-P flux in the stratosphere are preceded or accompanied by a similar event in the troposphere. Furthermore, this procedure has to be applied with caution, since the stratospheric flow is highly asymmetric during spring and some asymmetric patterns yield enhanced E-P flux that can be misinterpreted as the signature of bursts in planetary-wave activity (O'Neill and Pope, 1988). Randel (1987) and Randel et al. (1987) examined the time evolution of the amplitude and phase, as well as the E-P flux, of individual Fourier components at several levels in the troposphere and stratosphere during the SH winter. They found that the origin of planetary-wave activity in the stratosphere can, in some instances, be traced down to the troposphere. To our knowledge, therefore, this study represents the first attempt to apply techniques of nonlinear dynamics in an observational study of stratospheric phenomena.

The paper is organized as follows. The dataset used and the statistical methods applied are described in Section 2. In Section 3, EOF analyses of the flows in the stratosphere and troposphere of the SH during winter are used to describe the time evolution of these flows and the relation between these two evolutions. In Section 4, several case studies of enhanced planetary-scale activity in the SH troposphere-stratosphere during early winter are analyzed. A summary of results and conclusions are presented in Section 5.

## 2. Data and Statistical Methods

### a. *Data*

The dataset consists of the NMC daily 1200 UTC analysis fields of geopotential height at 18 levels between 1000 mb and 0.4 mb. At and below 100 mb, the data correspond to the operational analysis fields. Above 100 mb, the fields are produced using the retrievals from the stratospheric sounding units (SSU, MSU and HIRS-2) on board satellites of the TIROS-N series and the operational 100 mb analyses as base fields. Details of the data retrieval scheme are given by Gelman et al. (1987) and references therein.

The stratospheric data for the SH are not as reliable as those for the Northern Hemisphere (NH), partly because the 100 mb analysis fields in the SH rely on far fewer radiosonde measurements than those for the NH. However, it does not seem to make much difference whether the 100 mb analysis fields are taken from NMC, the United Kingdom Meteorological Office or the European Centre for Medium Range Weather Forecasts. The stratospheric fields obtained by using these three different analyses as the base fields were found to be in broad qualitative agreement (Workshop on the Intercomparison of Satellite Data from Different Sources, Williamsburg, VA; W.L. Grose and A. O'Neill, Eds.; in preparation).

### *b. EOF analysis*

We apply EOF analysis to the two time series of 10 mb and 500 mb geopotential height field anomalies in the SH covering seven (1979–1985) 92-day winter seasons (June-August). The fields are first transformed from the original $5° \times 2.5°$ longitude-latitude grid to the 380 point grid of Barnston and Livezey (1987), which has approximately equal distances between points. The seven-winter grand mean defined with respect to time is then subtracted from the daily fields yielding a set of daily anomalies at each grid point. Since we are interested primarily in those regions where the variability of the flow is largest, the anomalies are not normalized by the variance. The EOFs are the eigenvectors of the $380 \times 380$ covariance matrix thus obtained. The daily anomaly fields are then expanded in these EOFs to obtain the time coefficients, or principal components (PCs).

## 3. EOF Results

### *a. Stratosphere*

The first six EOFs of the time series of geopotential height fields at 10 mb are shown in Fig. 1. The corresponding variances associated with them are $7.2\pm1.1$, $5.9\pm0.8$, $3.6\pm0.5$, $3.4\pm0.4$, $3.2\pm0.4$ and $1.8\pm0.2\%$, respectively, of the total variance of the field. It is apparent that the first three EOFs are similar except for a zonal shift. They are all dominated by a high centered near 60°S and contain a large component with zonal wavenumber one. The high is located in the South Pacific, east of New Zealand, for EOF 1; directly south of Australia for EOF 2; and in the South Atlantic for EOF 3. EOFs 1 and 2 show weak lows almost 180° away from the corresponding highs.

EOFs 4 and 5 also have structures similar to each other, differing by a zonal shift. Both are dominated by a zonal wavenumber-two pattern with highs and lows centered slightly equatorward of 60°S. EOF 6 is much more zonally symmetric than the others, with a region of low values over Antarctica surrounded by a band of high values in mid latitudes. Its contribution to the total variance of the field is significantly less than any of the first five and so it will not be discussed further.

Degeneracy between EOFs 3, 4 and 5 appears to exist by examining their overlapping error bars. Mo and Ghil (1987) explained by simple examples why this is not a problem if the underlying dynamics can be seen to have distinct modes of variability with comparable variances. We shall see that the latter situation obtains by studying the behavior of the EOFs in time.

The fact that EOFs 1, 2, and 3–as well as 4 and 5–have similar patterns and differ mostly by a zonal shift suggests that they are associated with traveling, large-scale disturbances. An examination of the evolution of the PCs in individual winters lends support to this suggestion. The evolution of the PCs associated with the EOFs 1–3 during 1980 is shown in Fig. 2a. All three PCs show large values only during June and early July; the same is true in 1984 and 1985 (not shown). The three successive positive maxima in PCs 1–3 during June (first in PC 2, then in PC 1 and PC 3) are consistent with an eastward propagating disturbance traveling around a latitude circle. Aside from indicating a slower eastward movement, the evolution of the first three PCs during June 1985 is quite similar to that shown in Fig. 2a. In 1984 the large values are confined to PC 1 and PC 2, indicating the presence of a quasi-stationary disturbance to the flow.

The large values of PCs 1–3 during early winter 1982 (Fig. 2b) and 1979 (not shown) are mostly negative, corresponding to anomalies from climatology which are opposite in sign to those in 1980, 1984 and 1985. The magnitudes of the PCs 1–3, and hence the deviations from climatology, during 1981 (Fig. 2c) and 1983 (not shown) are remarkably small throughout almost the entire winter.

The behavior of PCs 4 and 5 in 1982 (Fig. 3) is the typical one: the largest values occur in late winter and individual maxima and minima in their amplitude alternate in time, suggesting again eastward moving disturbances. Their alternation is more rapid than that of PCs 1–3, indicating a faster eastward progression of these disturbances in which wavenumber two predominates.

The intervals of time when one of the five leading EOFs is a large and persistent contributor to the anomaly field are shown by the bars in Fig. 4. For each individual PC, bars are drawn when its magnitude exceeds one and one-half times its standard deviation in time during five or more consecutive days (the mean in time of each PC is zero). Filled and open bars represent large positive and negative values of the PCs, respectively.

The interannual variability in the winter circulation at 10 mb is concisely displayed in Fig. 4. Persistent events are more frequent and last longer in 1979, 1980, 1982, 1984 and 1985 than in 1981 and 1983. It is clear from Fig. 4 that the majority of the persistent events in EOFs 1–3 occur in early winter, while those in EOFs 4 and 5 occur in late winter. This and the different speeds of eastward propagation (Fig. 3 vs. Fig. 2) remove the suspicion of degeneracy raised by the overlapping error bars of the associated variances. The behavior of the first five PCs described here is consistent with observations showing that the zonal wavenumber one component of the flow in the stratosphere of the SH is characterized by largest amplitudes in early winter and spring, while the wavenumber-two component has largest amplitude during late winter and early spring (Geller and Wu, 1987; Randel, 1988).

Histograms of the individual PCs provide information, in a reasonably compact form, on the events that shape the variability of the circulation. The histograms, shown in Fig. 5, reveal that the amplitude of the first three PCs at 10 mb vary in a range that decreases as the order increases. Thus PC 1 is distributed between -5.5 km and 6.5 km, PC 2 between -4.5 km and 5.5 km, and PC 3 between -3.5 km and 4.5 km (with very few values outside this range). This aspect of their distribution hints at Gaussianity. For magnitudes $\leq 4$ km, however, the histogram for PC 1 shows a skewness toward negative values, primarily through the contributions of 1981, 1982 and 1983. In 1980, 1984 and 1985 PC 1 is more often positive than negative.

The histogram for PC 2 (Fig. 5b) shows strong skewness toward negative values at large magnitudes ($\geq 3$ km), but is skewed toward positive values at magnitudes smaller than 3 km. The values in 1982 are principally responsible for the negative skewness at large magnitudes. The large values in 1980, 1984 and 1985 are again mostly positive. In contrast to the first two PCs, a histogram for PC 3 (Fig. 5c) shows a slightly positively skewed distribution at magnitudes greater than 2 km (more so if we delete 1982), but is skewed toward negative values at smaller magnitudes. The large positive values occur again mostly in 1980, 1984 and 1985. Histograms for PCs 4 and 5 (not shown) are more symmetric, probably as a result of the more rapid eastward progression of the patterns associated with them, as noted above.

*b. Troposphere*

The first two EOFs of the time series of the 500 mb geopotential height anomalies in our dataset closely resemble the first and third EOFs obtained by Mo and Ghil (1987) using a different dataset. They are associated with 12.6 and 9.1%, respectively, of the total variance of the field. EOF 1 is essentially zonally symmetric with anomalies in mid latitudes out of phase with those in high latitudes. The corresponding seesaw in height anomalies between middle and high latitudes is closely tied to fluctuations in the strength of the westerly flow south of 40°S. The variation in time of this EOF 1 pattern is mostly on interannual timescales and reflects the interannual variability in the SH winter zonal mean flow (van Loon, 1972; Trenberth, 1979). Our EOF 2 is the same as Mo and Ghil's EOF 3 (their Fig. 12c), called the Pacific-South American (PSA) pattern because it is nearly a mirror image of the Pacific-North American (PNA) pattern (Horel and Wallace, 1981).

Our EOF 3 (Fig. 6), accounting for 8.1% of the variance, does not resemble any of the previously published eigenvectors of the SH 500 mb height field. Its structure is almost identical to the composite presented by Mo (1986, Fig. 8b) of four quasi-stationary events of 500 mb geopotential anomalies dominated by a zonal wavenumber 4 pattern. This discrepancy between EOFs is a consequence of a slight difference in the analysis procedure. Unlike our EOF analysis, Mo and Ghil (1987) retained only zonal wavenumbers zero through four in the geopotential field. This accounts for the absence of a wavenumber four-type pattern of EOF from their set of EOFs (Farrara et al., 1988). Our EOF 4, which explains 7.4% of the variance of the field, corresponds to EOF 2 in Mo and Ghil's (1987; their Fig. 12b) analysis (with a change in sign which is immaterial) and is similar to a composite of zonal wavenumber-three quasi-stationary events in Mo (1986, her Fig. 9). After EOF 4 there is a sharper dropoff in explained variance for the higher-order EOFs, so we will restrict our discussion of the tropospheric EOFs to the first four.

The persistent events in the troposphere, as seen through our first four EOFs, are displayed graphically in Fig. 7. As in Fig. 4, the bars represent the intervals of time when the magnitude of the individual PCs exceed one and one-half times their standard deviation in time during five or more consecutive days. This criterion results in the identification of a greater number of persistent events in the SH troposphere than the pattern correlation criterion of Horel (1985), Mo (1986) and Mo and Ghil (1987). During the years 1979–1982, when our data overlap, each of the quasi-stationary events identified by Mo (1986) also appears as a persistent event in Fig. 7.

Careful inspection of Figs. 4 and 7 suggests that there are no clear connections between tropospheric and stratospheric persistent anomalies. For instance, 1985 is the most quiescent year of the set in the troposphere, while it is one of the more disturbed years in the stratosphere. Simultaneous or time-lagged correlations between PCs at 500 mb and 10 mb, do not exhibit significant values either. This lack of systematic association, from the point of view of the persistent anomalies, between the troposphere and the stratosphere suggests that the low-frequency variability in one region is independent of the other. In particular, the planetary-scale structures that dominate the stratosphere of the SH during winter do not seem to be connected with persistent quasi-stationary events in the troposphere of either synoptic or planetary scale.

The only troposphere-stratosphere pair that shows any indication of being correlated is PC 1 at 500 mb and PC 1 at 10 mb. The scatter diagram in Fig. 8 shows that the two PCs are positively correlated at large magnitudes. In other words, when geopotential heights at 500 mb are lower (higher) than normal over Antarctica those at 10 mb over the South Pacific tend to be lower (higher) than normal. Therefore, during winters with stronger (weaker) than normal mid-latitude westerlies at 500 mb the stratospheric vortex at 10 mb tends to be displaced toward the South Pacific (South Atlantic).

## 4. Case studies

In this section, we examine selected early winter periods in greater detail and attempt to relate the evolution of the flow in the stratosphere to that in the troposphere by the more conventional methods of Fourier decomposition. The selected periods correspond to years with strong disturbances in the SH stratosphere, on the one hand, and to years when the stratospheric circulation was weakly disturbed on the other.

### a. *Traveling wavenumber one*

Maps of the deviations from the zonal mean of the geopotential height field at 10 mb for a series of days during June 1980 are shown in Fig. 9. All of these maps show a predominantly zonal wavenumber one-type pattern. On 1 June (Fig. 9a) the high of the pattern is strong and located south of Australia slightly poleward of 60°S. The high intensifies as it moves eastwards across the South Pacific and slowly decays while continuing to move steadily eastward into the South Atlantic (Figs. 9b, c and d). During this period, therefore, the cyclonic circumpolar vortex in the middle stratosphere became clearly displaced towards the South Atlantic, particularly in early June.

Maps of the geopotential height anomalies for the same four days (Figs. 10a-d) strongly resemble the maps in Fig. 9, except that the negative values are much weaker. It is apparent from Figs. 10 and 1 that the eastward-traveling zonal wavenumber one-type disturbance in Fig. 9 is associated with the three large peaks in PCs 1–3 in Fig. 2a. A set of similar maps for a series of days in June 1985 (not shown) shows a very similar sequence of events, with a strong zonal wavenumber one-type disturbance showing the same geographical preference for development, translation and decay, but somewhat slower eastward movement.

Figure 11 shows a longitude-height section of the geopotential height deviations averaged over latitudes 50–60°S throughout the troposphere and stratosphere for the same four days in 1980 as in Fig. 9. On 1 June a region of positive values extends upward and westward from the middle troposphere near 180° to cover the entire stratosphere of the eastern hemisphere. Another region of positive values around 120°W is confined to the troposphere. The negative values in the stratosphere of the western hemisphere are less organized than the positive values in the eastern hemisphere. On 8 June the deviations show a strong zonal wavenumber-one pattern, which is nearly equivalent barotropic throughout the stratosphere (Fig. 11b). The region of large positive values in the stratosphere has moved from around 60°E (Fig. 11a) to near the dateline and appears no longer connected with similar regions in the troposphere. During the following 8 days the pattern in the middle stratosphere continues to move eastward and decrease in amplitude (Fig. 11c) and by 24 June it has returned to nearly the same configuration as on 1 June (Fig. 11d).

The evolution at 500 mb of the component of the geopotential height field with zonal wavenumber one during June 1980 is shown in Fig. 12. At this level, the wave is quasi-stationary throughout June, with its trough near 30°E, and has relatively small amplitude after an amplification early in the month. Similar plots for 1985 (not shown) reveal a strikingly similar sequence of events during June: the growth of a strong zonal wavenumber one-type disturbance in the stratosphere early in the month, followed by steady eastward movement and decay. These stratospheric events are preceded by an amplification of a quasi-stationary wave with zonal wavenumber one in the middle troposphere, although the trough of this wave is near 60°E in 1985.

In 1985, once the wavenumber one-type disturbance in the stratosphere is strong, the disturbance develops a second, weaker region of positive deviations 180° away from the stronger, primary area of positive values. The pattern of the deviations from the zonal mean of the geopotential height field on 18 June 1985, a date which corresponds approximately to that of Fig. 9c in the evolution of the 1980 disturbance, is shown in Fig. 13a. The secondary region of positive values is located south of South Africa and the negative values have split into two centers.

These developments are reflected in a persistent event in EOF 4 during the latter half of June 1985 (Fig. 4) and in the increase in amplitude of the zonal wavenumber two component of the flow and the decrease in that of zonal wavenumber one (not shown). Indirect evidence that such behavior involves nonlinear coupling of zonal wavenumbers 1 and 2 comes from the study of O'Neill and Pope (1988). They found a similar sequence of events in idealized numerical experiments of strong disturbances in the stratosphere, and presented evidence that the waves were nonlinearly coupled in the simulations. The amplitudes of the waves studied here are comparable to those in the simulations.

A possible scenario for the events during June of 1980 and 1985 is that the zonal wavenumber one-type stratospheric disturbances are triggered by the amplification of zonal wavenumber one in the troposphere, but then evolve independently of events in the troposphere at later times.

During early winter of 1982, PC 1 has large magnitude with negative sign at both

500 mb and 10 mb. The eddy (i.e., the deviation from the zonal mean) and anomaly components of the geopotential height at 10 mb on 12 June 1982, a day when the flow was typical of early winter in that year, are shown in Fig. 14. The eddy component (Fig. 14a) is weak, although its pattern resembles that of the strong disturbances in 1980 and 1985 at their peak amplitude. The corresponding anomalies (Fig. 14b) are almost all negative at high latitudes, with strong negative values in the South Pacific and weak positive values in the South Atlantic.

The early winter circulation during 1979 is similar to that in 1982, being characterized by a deeper than normal polar vortex from the troposphere to the middle stratosphere and weak wave activity in the stratosphere. Interestingly the zonal wavenumber one component of the geopotential height field at 500 mb has below normal amplitudes during early winter in both 1979 and 1982. This is consistent with the findings of Trenberth's (1980) study of planetary waves at 500 mb in the SH. Trenberth showed that if the westerlies in mid-latitudes are stronger than normal the amplitude of the wavenumber one component of the flow is weaker than normal.

*b. Quasi-stationary wavenumber one*

During 1984, in contrast to 1980 and 1985, a large zonal wavenumber one-type disturbance develops in the stratosphere during late June/early July (Fig. 15a) without being preceded by an amplification of the zonal wavenumber one component of the flow in the troposphere (Fig. 15b). Enhancing the contrast, the stratospheric disturbance in 1984 remains quasi-stationary.

During 1981 and 1983, the behaviors of the zonal wavenumber-one component of the flow in the troposphere and stratosphere seem reversed compared to 1984. The amplitude of this component at 500 mb during winter 1983 is shown in Fig. 16a. During early June 1983, its amplitude is large and the phase is approximately the same as in 1980 and 1985; still, by 10 June of that year no substantial large-scale disturbance has developed in the stratosphere (Fig. 16b). In addition, there are no subsequent developments, despite the appearance of the same vertically extended-structures in the eddy component on the longitude-height plane as in 1980 (Fig. 11a) and 1985. Similar remarks can be made about the early winter circulation in 1981.

Planetary-scale waves might have propagated more readily in the vertical during 1980 and 1985 than in 1981 and 1983. To verify this we inspected the variations in the quasi-geostrophic refractive index squared (Mechoso et al., 1985), which represents the effect of the mean zonal flow on the propagation of a steady wave in the framework of quasi-geostrophic linear wave theory. But, very little difference exists between the mean zonal flows in the lower stratosphere, and hence in the refractive index, in 1981/1983 and 1980/1985. In fact, the flow in June is never favorable, from the linear point of view, for the upward propagation of steady planetary waves into the stratosphere.

## 5. Summary and Discussion

We studied the observed variability of the SH stratospheric circulation during winter. This variability was described statistically in terms of the six leading EOFs of the 10 mb geopotential height field anomalies. Although the stratospheric dataset is too short (seven years) to allow a rigorous statistical analysis, it is sufficiently long to provide a broad picture of the events that shape the variability of the stratospheric circulation. The first three EOFs of the 10 mb geopotential height fields all have a similar structure and are associated with large, sometimes eastward-traveling, wavenumber one-type disturbances. These disturbances develop in early winter of some years (1980, 1984, 1985) over a preferred geographical location (south of Australia). The fourth and fifth EOFs in the stratosphere are associated with eastward-moving zonal wavenumber two-type disturbances, which are generally active in late winter.

In order to relate the variability in the stratosphere to that in the troposphere we also performed an EOF analysis of 500 mb geopotential height field anomalies. Three (the first, second and fourth) of our leading four tropospheric EOFs correspond to EOFs obtained by Mo and Ghil (1987) using a different dataset. Although our third EOF has no counterpart among the previously published EOFs of the SH 500 mb height field, it has the same structure as a composite of wavenumber four dominated quasi-stationary events presented by Mo (1986). This reinforces the evidence presented by Ghil (1987, 1988) and by Mo and Ghil (1987) for the connection between EOFs and persistent anomalies (Farrara et al., 1988). There are no apparent relationships between the time variations in these first four EOFs in the troposphere and those of the first five EOFs in the stratosphere. Variations, primarily interannual, in the sign of PC 1 at 500 mb are associated with variations in the strength of the westerly vortex throughout the troposphere and stratosphere, resulting in a weak correlation between PC 1 at 500 mb and PC 1 at 10 mb.

An examination of the individual early winter episodes of enhanced wave activity, however, reveals that, in some cases (1980, 1985), the growth of the zonal wavenumber one-type disturbances in the stratosphere is preceded by an amplification of the zonal wavenumber-one component of the flow in the troposphere. In these cases, the stratospheric disturbance that has developed moves subsequently eastward, while the tropospheric wave remains quasi-stationary. In 1984 the stratospheric disturbances are strong and quasi-stationary during late June and early July, though the tropospheric zonal wavenumber-one component of the flow remains weak throughout the first half of the winter.

During early winter of 1983 and 1981, the disturbances in the stratosphere are quite weak, despite the sustained activity of the tropospheric wavenumber one (with the same phase as that in 1980 and 1985). During early winter of 1979 and 1982, on the other hand, the circulation is characterized by a stronger-than-normal westerly vortex and weaker-than-normal zonal wavenumber one activity throughout the troposphere and stratosphere. The large eastward-moving zonal wavenumber one-type disturbances that sometimes develop in the early winter of the SH are responsible for the early winter maxima in wavenumber one amplitudes noted by Geller and Wu (1987) and Randel (1988).

This study shows that, the closer we look at the SH stratospheric circulation, the farther our view departs from that of an essentially-zonal flow perturbed by quasi-linear planetary-scale waves extending from the troposphere. We cannot find a clear relation between the generation and evolution of large-scale disturbances in the SH stratosphere and tropospheric blocking. This statement holds, even and especially, for early winter, when the zonal mean circulation is strongest in the seasonal cycle, and closest agreement with linear wave theory may be expected.

At best, we find a weak relation between the planetary-scale disturbances in the stratosphere and planetary-scale waves in the troposphere. The relation is blurred, however, during the subsequent evolution of the two phenomena, which suggests an increasing role for local nonlinearities. Systematic features of the wave development and evolution are linked to the orography.

Are these early-winter disturbances related in any way related to the annual recurring spring 'ozone hole' over Antarctica? Both phenomena seem to be related to the quasi-biennial oscillation (QBO) in the lower stratospheric equatorial zonal flow. The early-winter disturbances studied here develop during the 'westerly phase' of the QBO, when the zonal flow at this location is westerly. The 'ozone hole' is generally deeper in years when the QBO is in its westerly phase than in its easterly phase (Garcia and Solomon, 1987).

A possible hypothesis for this relation might develop along the following lines. In years when the QBO is in its westerly phase, the stratospheric circulation is preconditioned to amplify the tropospheric forcing. The resulting stratospheric disturbances cause asymmetries in the polar night jet. In this way, some parcels of air that would otherwise not be insolated would, on part of their trajectories around the pole, be affected by solar radiation. This radiation would affect their chemical composition, particularly the type of chlorine compounds they carry. These parcels will diffuse towards the pole and the lower stratosphere throughout the winter and affect the polar lower stratosphere in spring.

*Acknowledgements.* We thank Dr. Kingtse C. Mo for providing the EOF analyses programs. Support for this work was provided by NASA under Grants NAGW-1021 and NAG-5713, and by NSF under Grant ATM86-15424.

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## Figure Captions

Fig. 1 First six EOFs (empirical orthogonal functions) of the time series of of 10 mb winter (June-August) geopotential height anomalies. EOFs are normalized to unit vector length $\times$ 100.

Fig. 2 Amplitude (km) of the first three principal components during winter at 10 mb (PC 1 solid, PC 2 short dashes, PC 3 long dashes) for (a) 1980, (b) 1982 and (c) 1981.

Fig. 3 Amplitude (km) of the fourth (solid) and fifth (dashed) principal components at 10 mb during winter 1982.

Fig. 4 Persistent events (solid bars for positive values of the PCs, hatched for negative; see text for explanation) in each of the first five EOFs at 10 mb for the years 1979-1985.

Fig. 5 Histogram of PC amplitude (km) at 10 mb for (a) PC 1, (b) PC 2 and (c) PC 3. Solid curve shows a Gaussian distribution with the same mean and variance as each PC.

Fig. 6 Third EOF of the time series of 500 mb winter geopotential height anomalies. EOFs normalized to unit vector length $\times$ 100.

Fig. 7 As in Fig. 4, except for the first four EOFs at 500 mb.

Fig. 8 Scatter diagram of the amplitude (m) of PC 1 at 10 mb vs. the amplitude of PC 1 at 500 mb. Values are plotted for each day of the 644 day time series. The numbers denote the year in which each particular point occurred (e.g., 9=1979, 0=1980, 1=1981, etc.). The box encloses the region where the amplitudes at both levels are less than one and one-half times their standard deviation in time.

Fig. 9 Deviation from the zonal mean of geopotential height (m) at 10 mb for (a) 1 June, (b) 8 June, (c) 16 June and (d) 24 June 1980.

Fig. 10 As in Fig. 9, except for the anomaly of geopotential height (m) at 10 mb from its climatological mean.

Fig. 11 Deviation from the zonal mean of geopotential height (m), averaged over the latitudes 50-60$^0$S, as a function of height and longitude for (a) 1 June, (b) 8 June, (c) 16 June and (d) 24 June 1980.

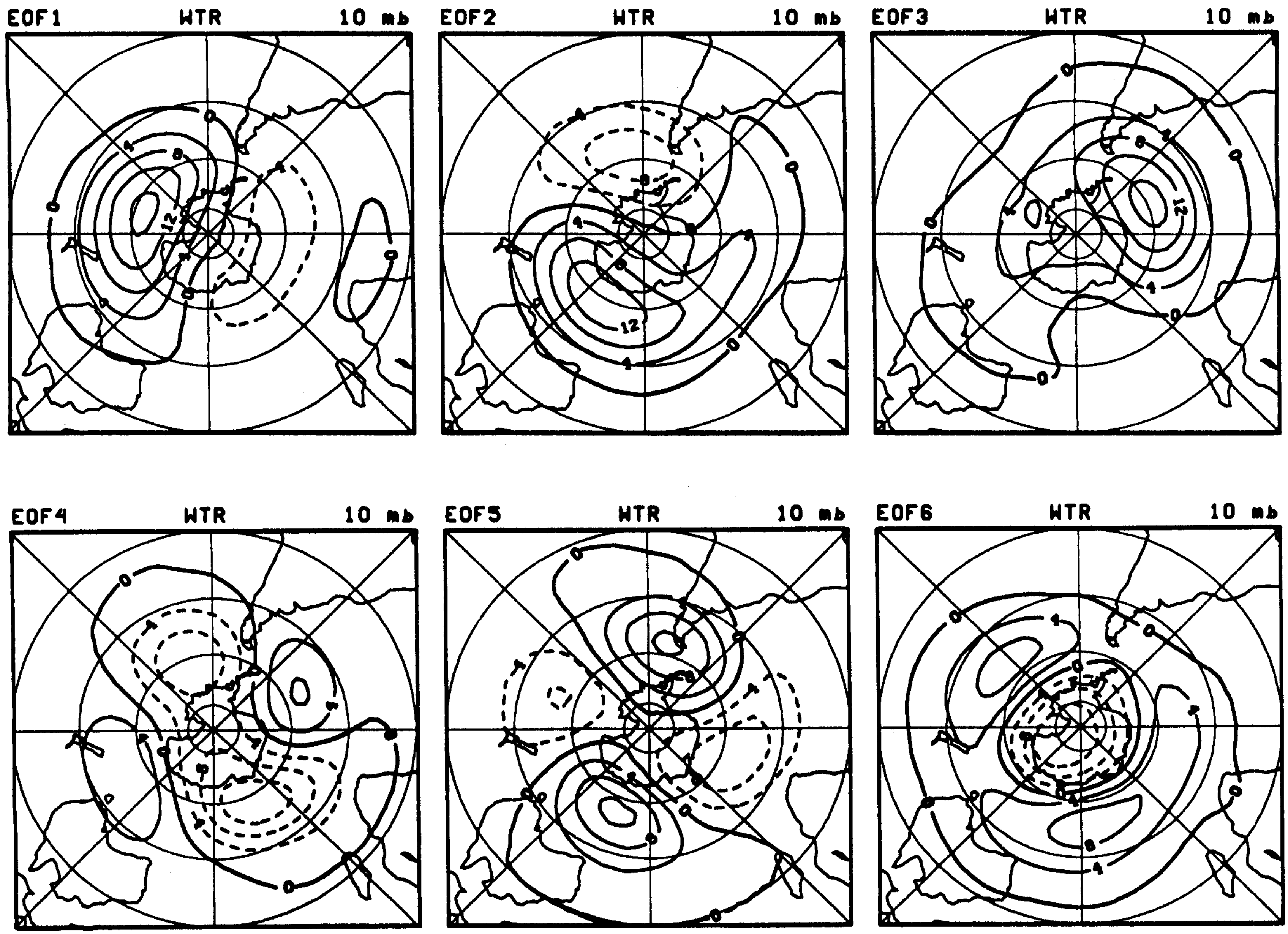
Fig. 12 Wavenumber one component (m) of the geopotential height field at 500 mb averaged over the latitudes 50-60$^0$S during June 1980.

Fig. 13 Deviation from the zonal mean of geopotential height (m) at 10 mb for 18 June 1985.

Fig. 14 Deviation of geopotential height (m) at 10 mb from (a) its zonal mean and (b) its climatological mean, for 12 June 1982.

Fig. 15 As in Fig. 12 except for (a) 10 mb and (b) 500 mb during winter 1984.

Fig. 16 (a) As in Fig. 12 except for winter 1983 and (b) as in Fig. 11 except for 10 June 1983.



EOF1
WTR
10 mb
EOF2
WTR
10 mb
EOF3
WTR
10 mb
EOF4
WTR
10 mb
EOF5
WTR
10 mb
EOF6
WTR
10 mb


Figure 1

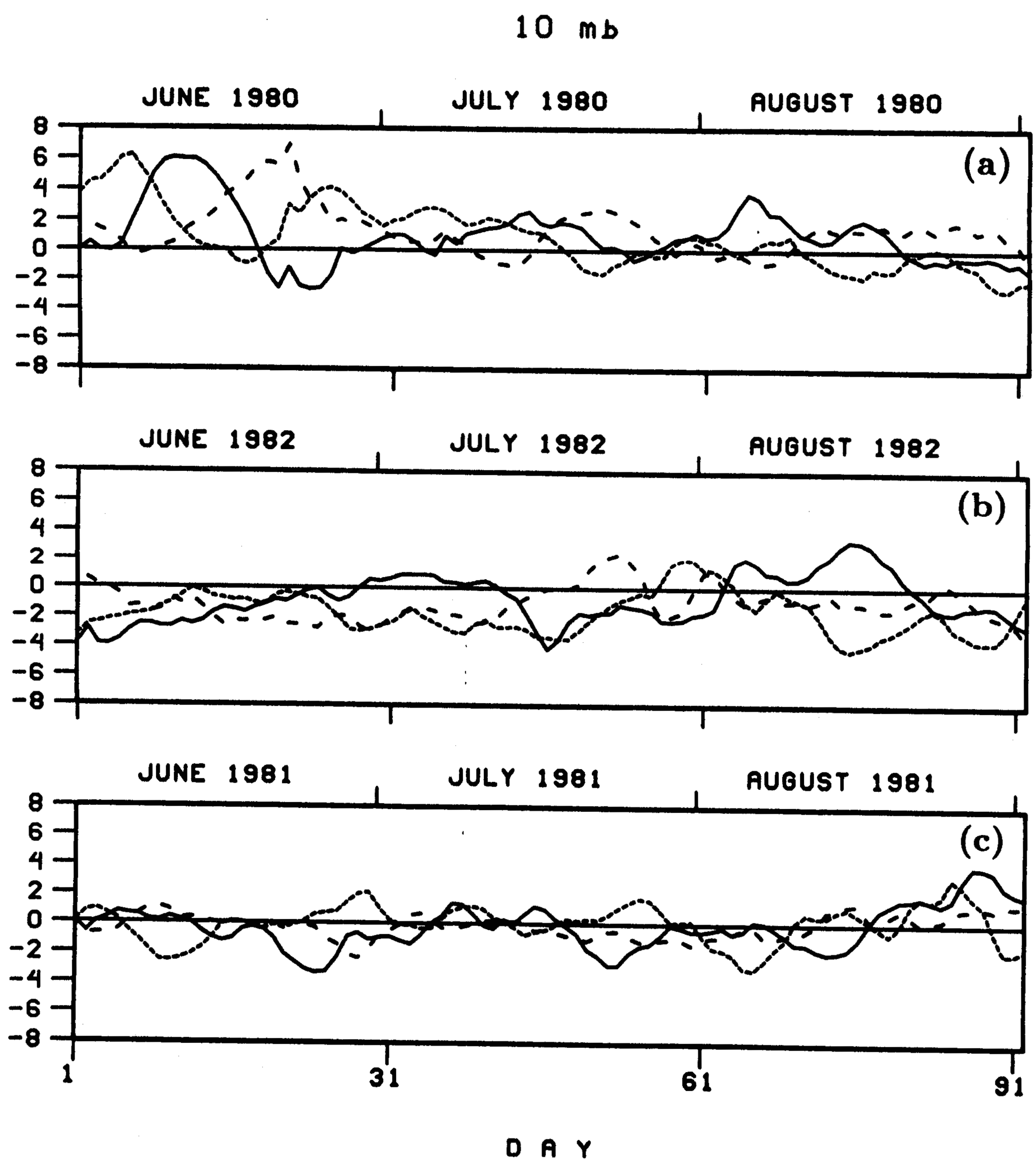

10 mb
JUNE 1980
JULY 1980
AUGUST 1980
(a)
JUNE 1982
JULY 1982
AUGUST 1982
(b)
JUNE 1981
JULY 1981
AUGUST 1981
(c)
8
6
4
2
0
-2
-4
-6
-8
1
31
61
91
DAY

**Figure 2**

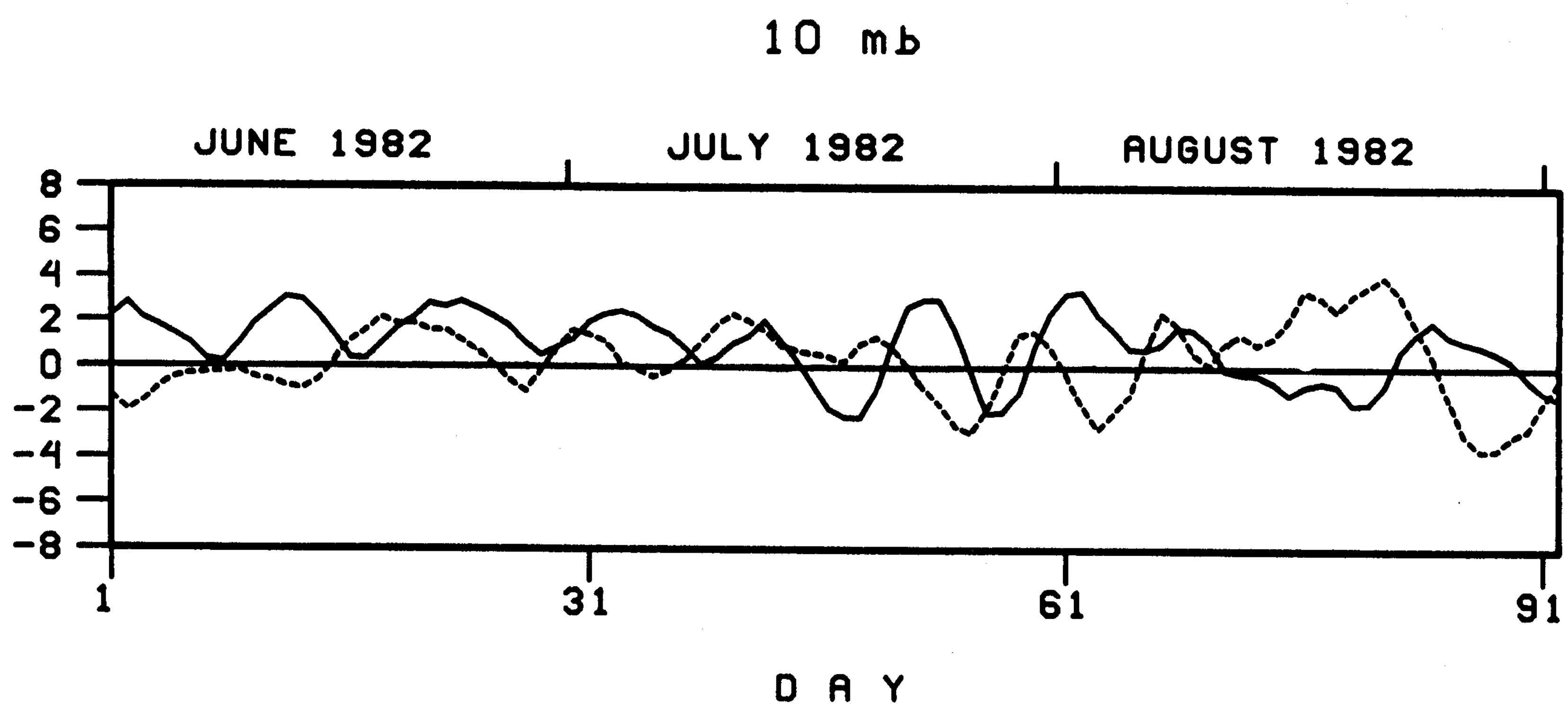



**Figure 3**

## 10 mb

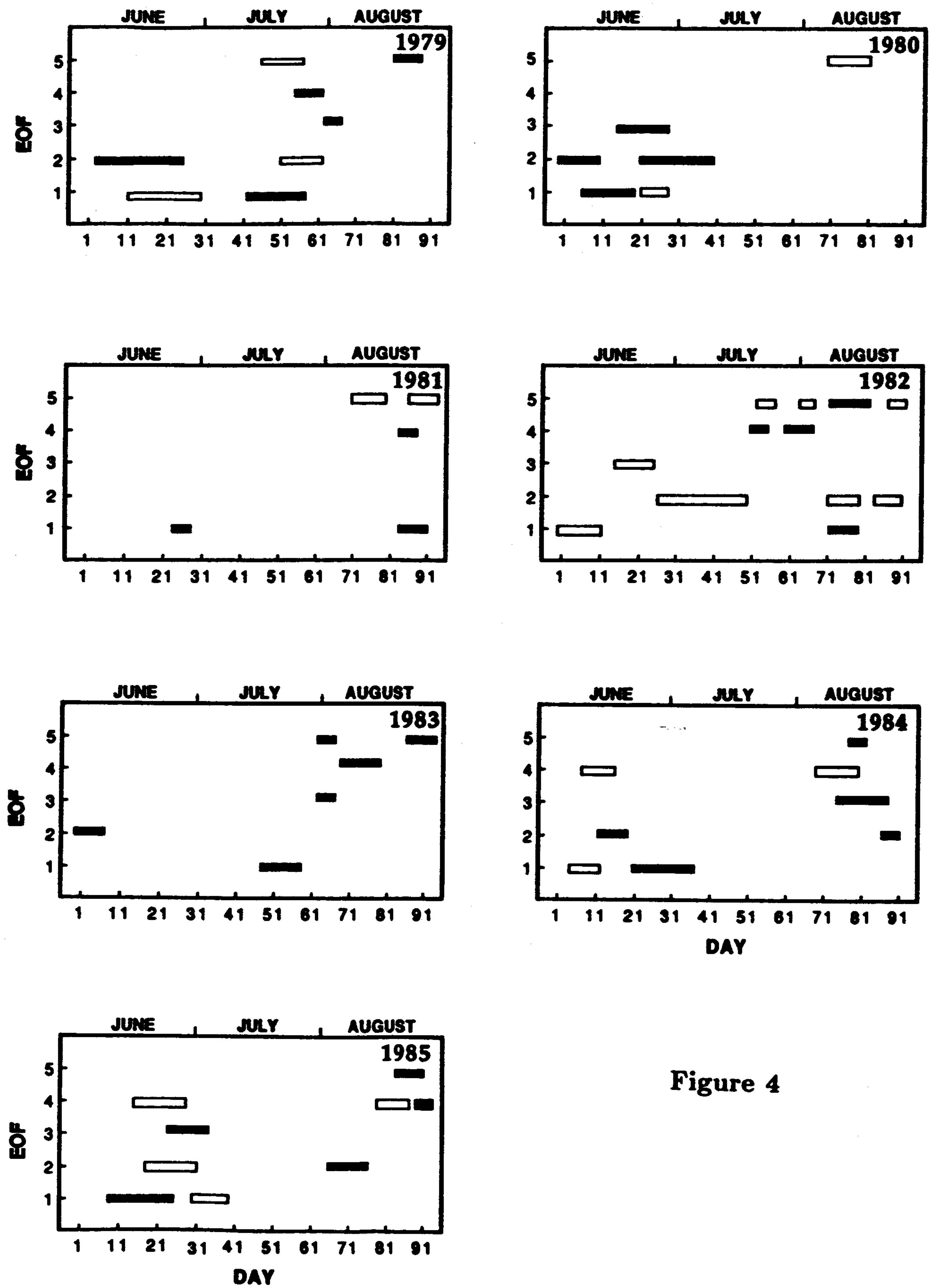



Figure 4

(a) PC-1 10 mb

(b) PC-2 10 mb

(c) PC-3 10 mb

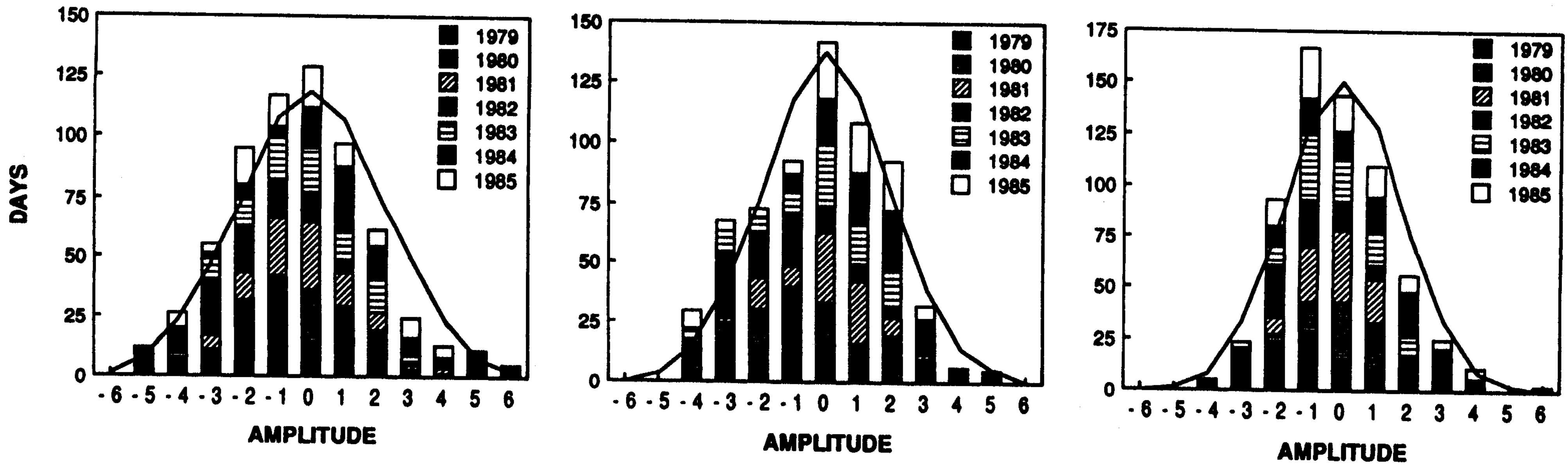



Figure 5

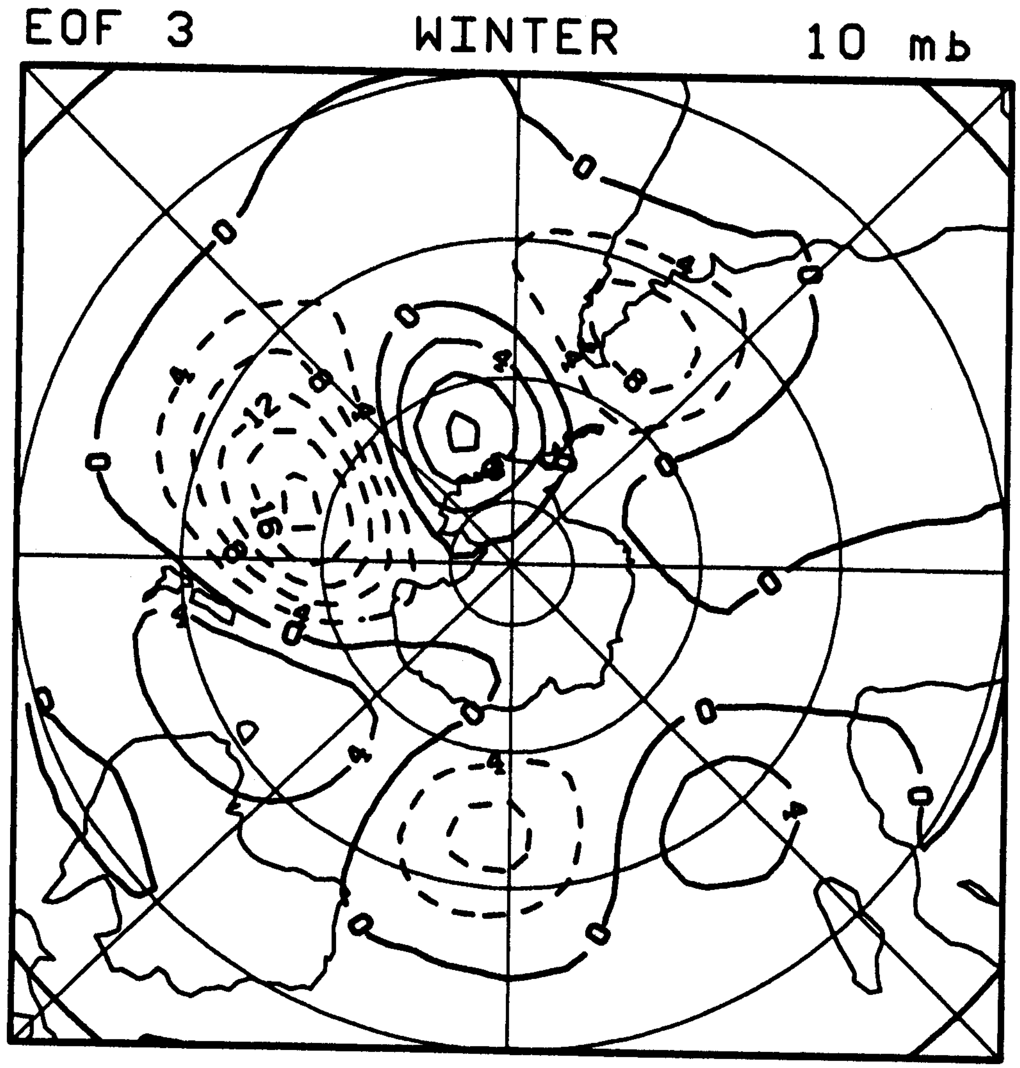



**Figure 6**

**500 mb**

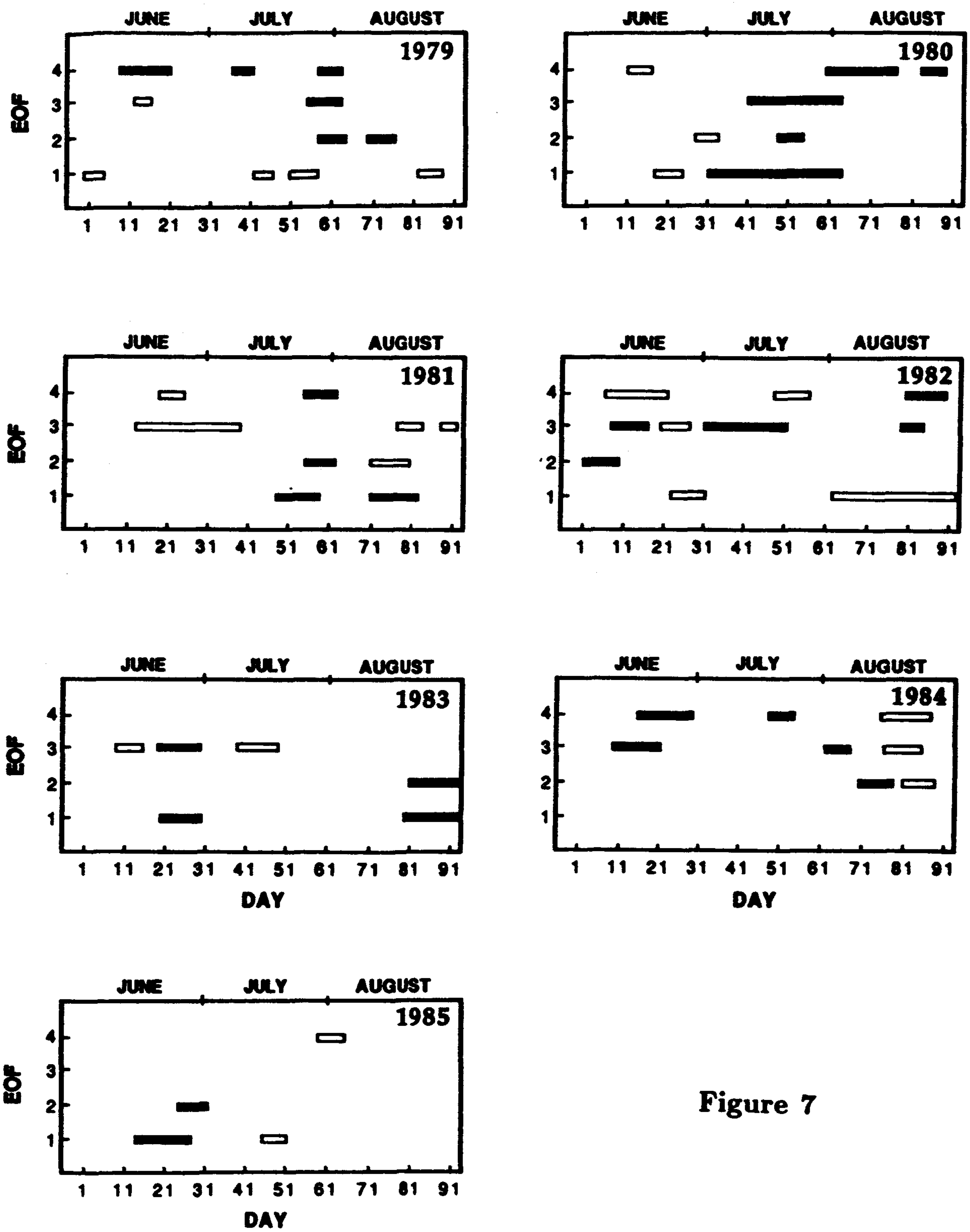



Figure 7

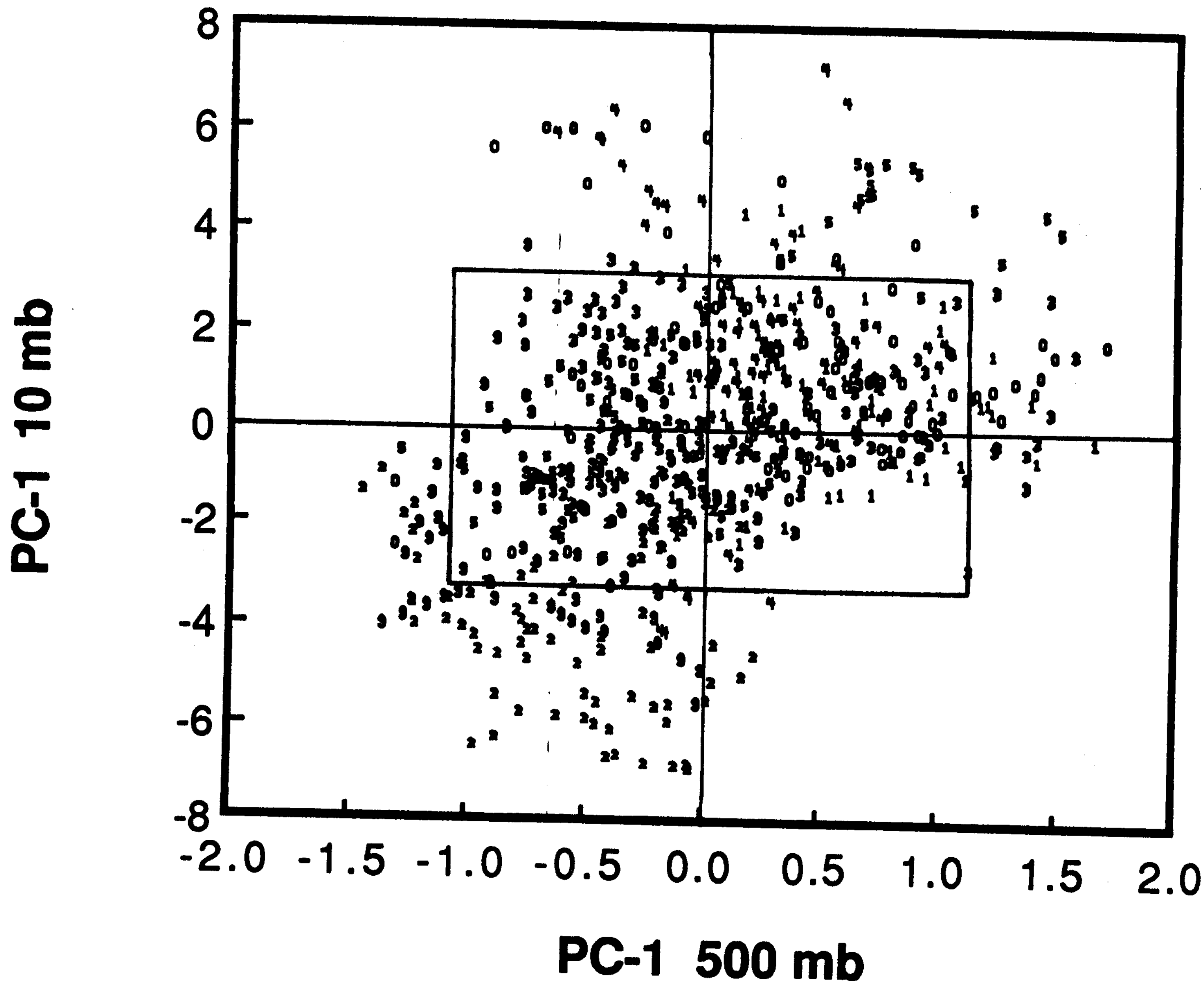



Figure 8

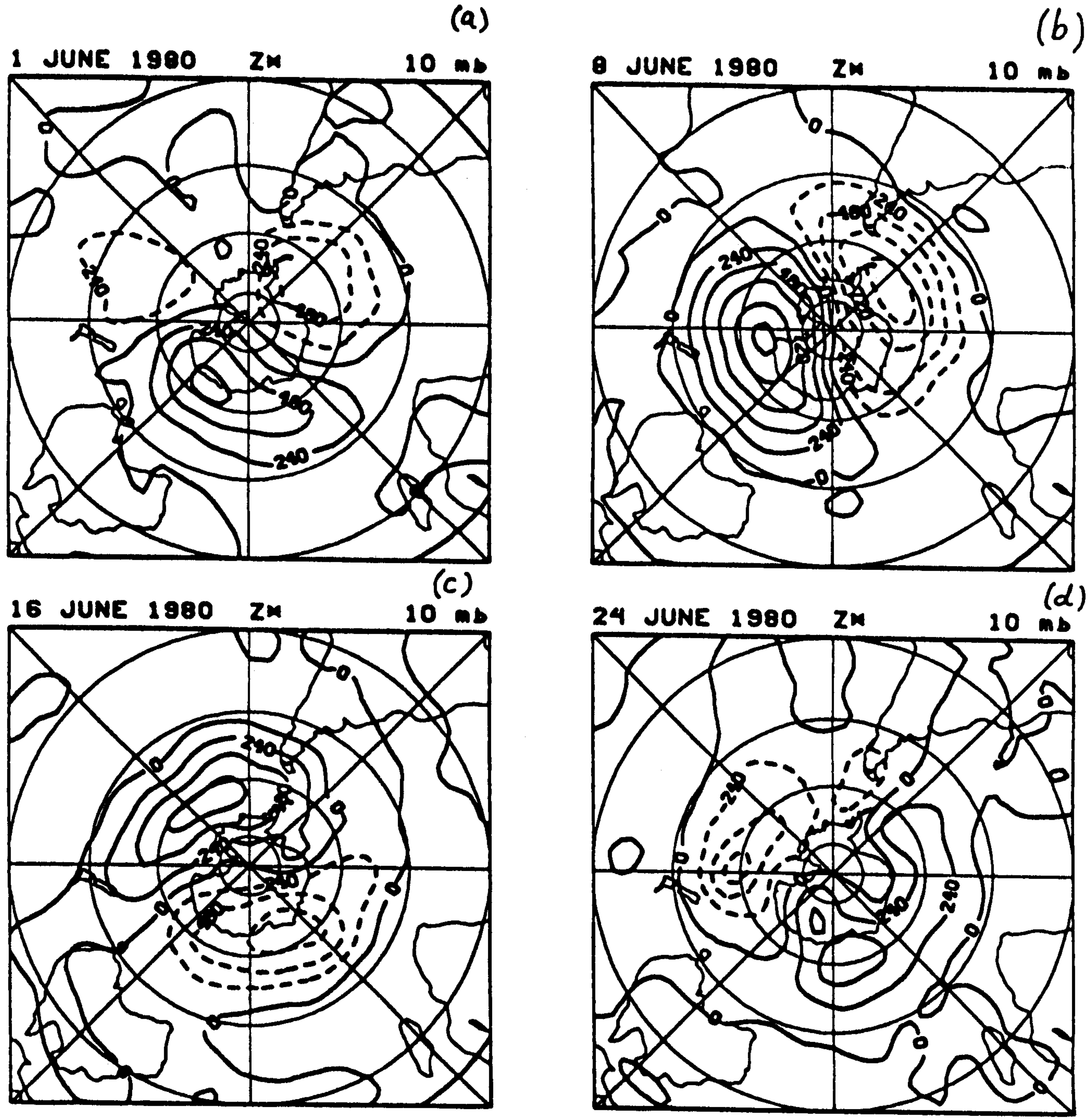

(a)
1 JUNE 1980 ZM 10 mb
(b)
8 JUNE 1980 ZM 10 mb
(c)
16 JUNE 1980 ZM 10 mb
(d)
24 JUNE 1980 ZM 10 mb

Fig. 9

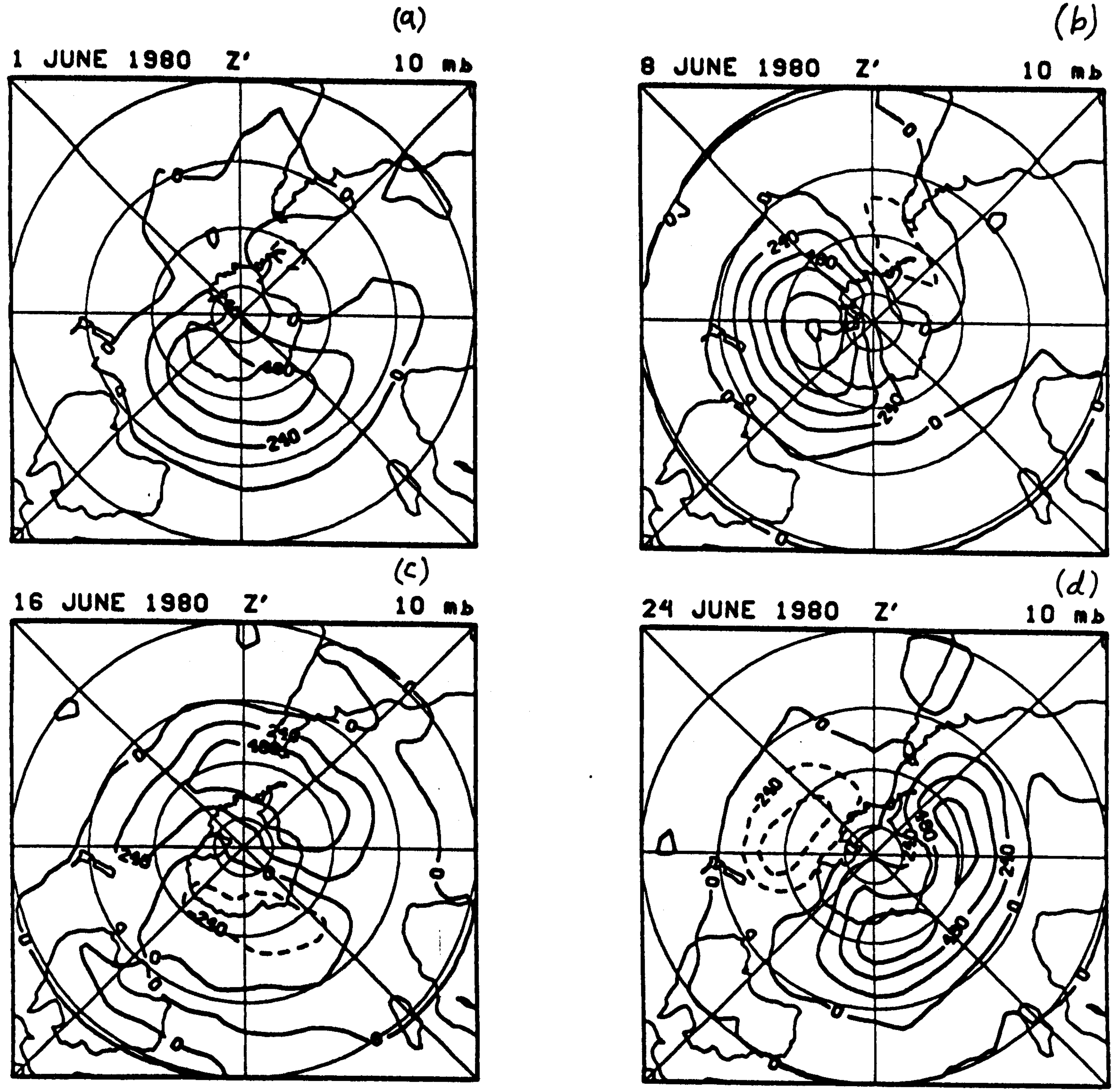

(a)
1 JUNE 1980 Z' 10 mb
0
240
480
(b)
8 JUNE 1980 Z' 10 mb
0
240
480
(c)
16 JUNE 1980 Z' 10 mb
0
240
480
-240
(d)
24 JUNE 1980 Z' 10 mb
0
240
480
-240

Fig. 10

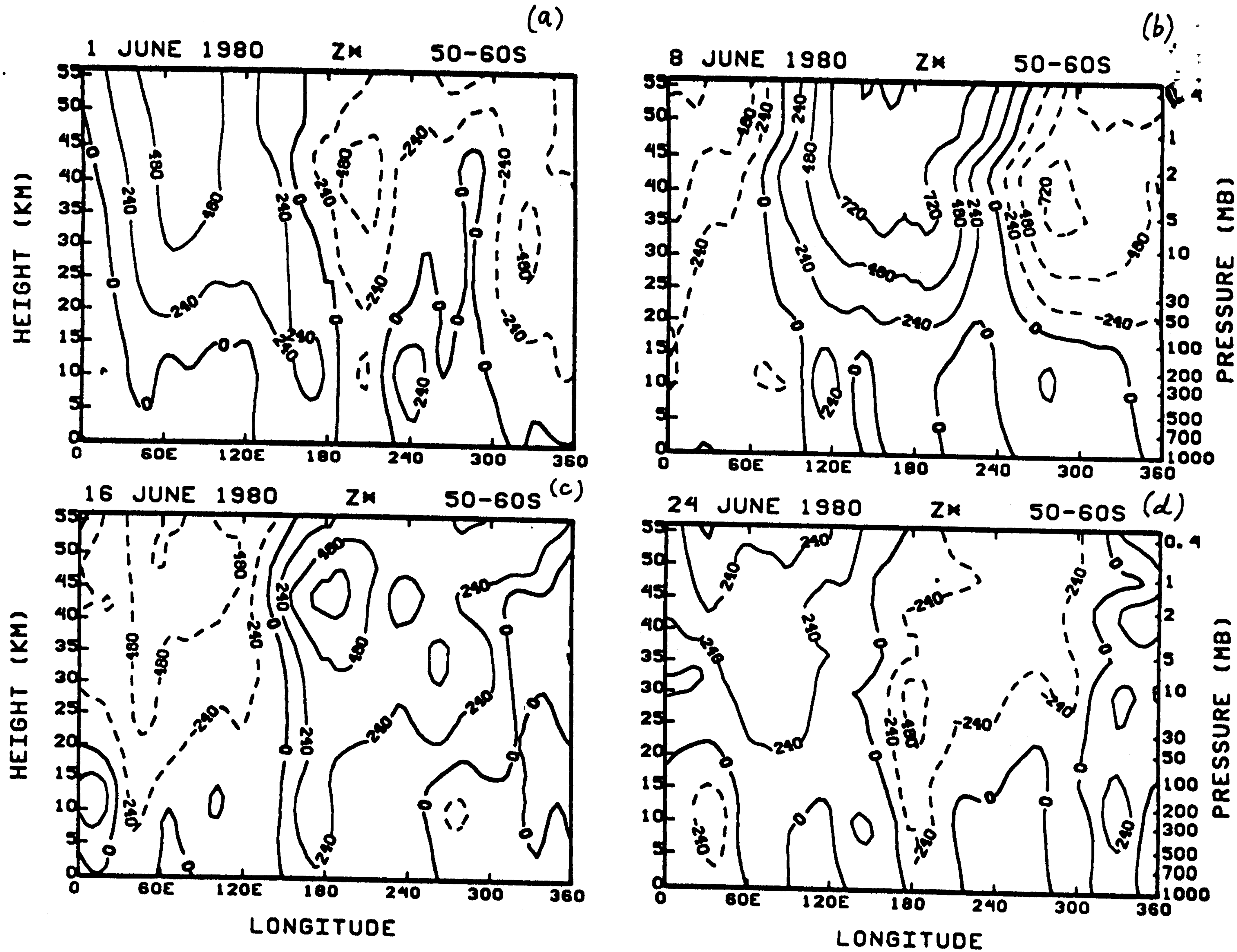
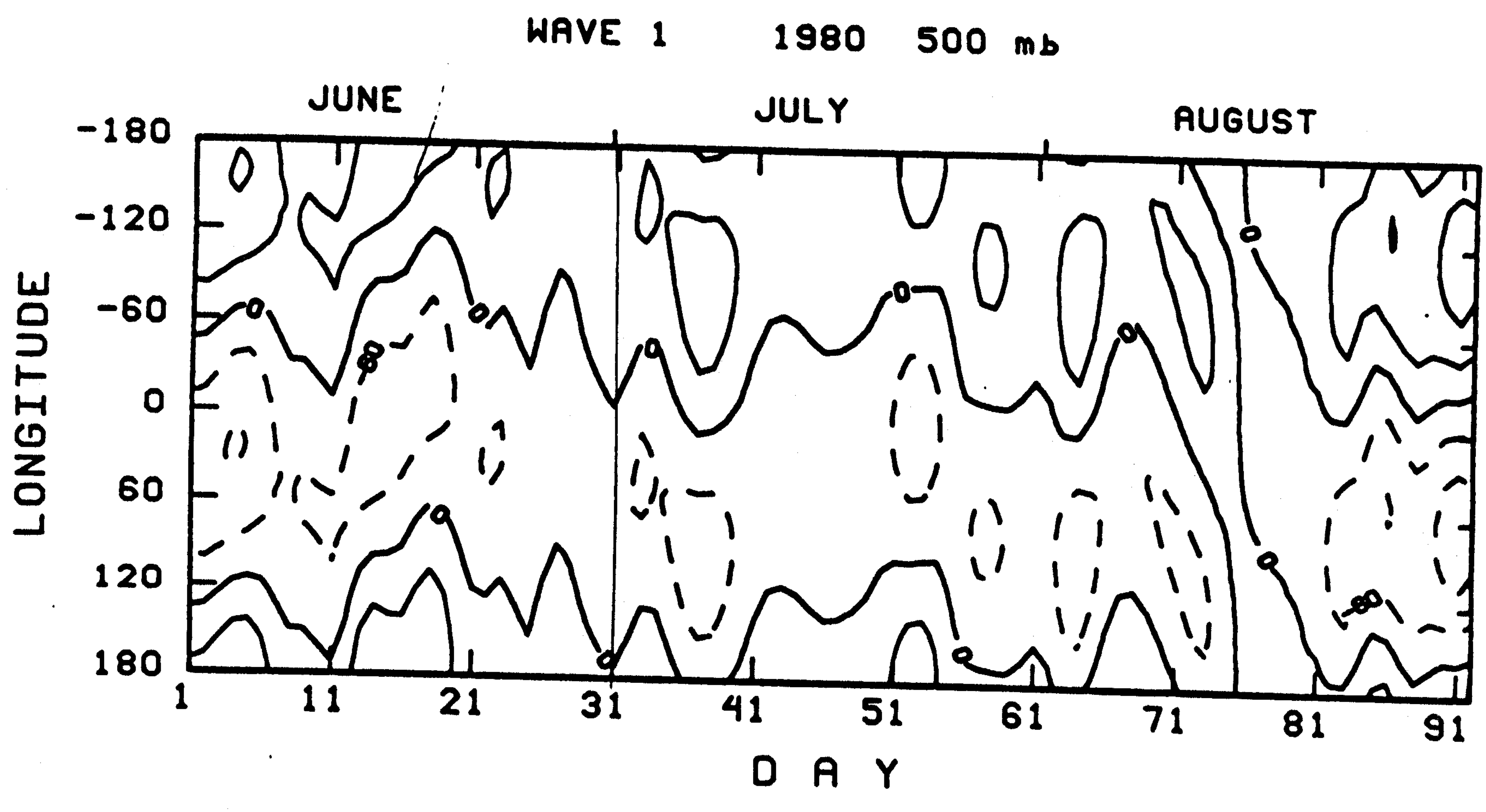

(a)
1 JUNE 1980 Z* 50-60S
(b)
8 JUNE 1980 Z* 50-60S
(c)
16 JUNE 1980 Z* 50-60S
(d)
24 JUNE 1980 Z* 50-60S
HEIGHT (KM)
PRESSURE (MB)
LONGITUDE

Fig. 11

60 ?



WAVE 1 1980 500 mb
JUNE
JULY
AUGUST
LONGITUDE
-180
-120
-60
0
60
120
180
1
11
21
31
41
51
61
71
81
91
DAY


Fig. 12

18 JUNE 1985 Z* 10 mb

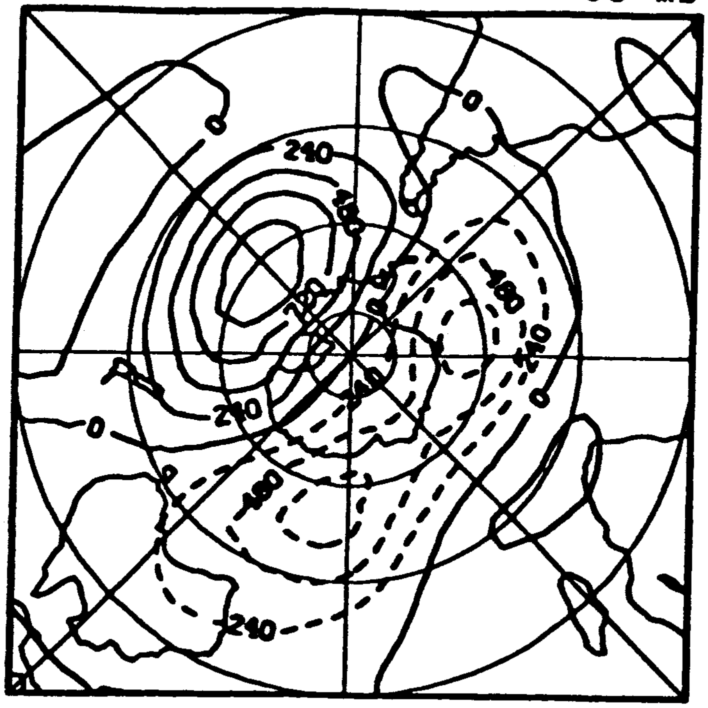



Figure 13

(a)

12 JUNE 1982 ZM 10 mb

(b)

12 JUNE 1982 Z' 10 mb

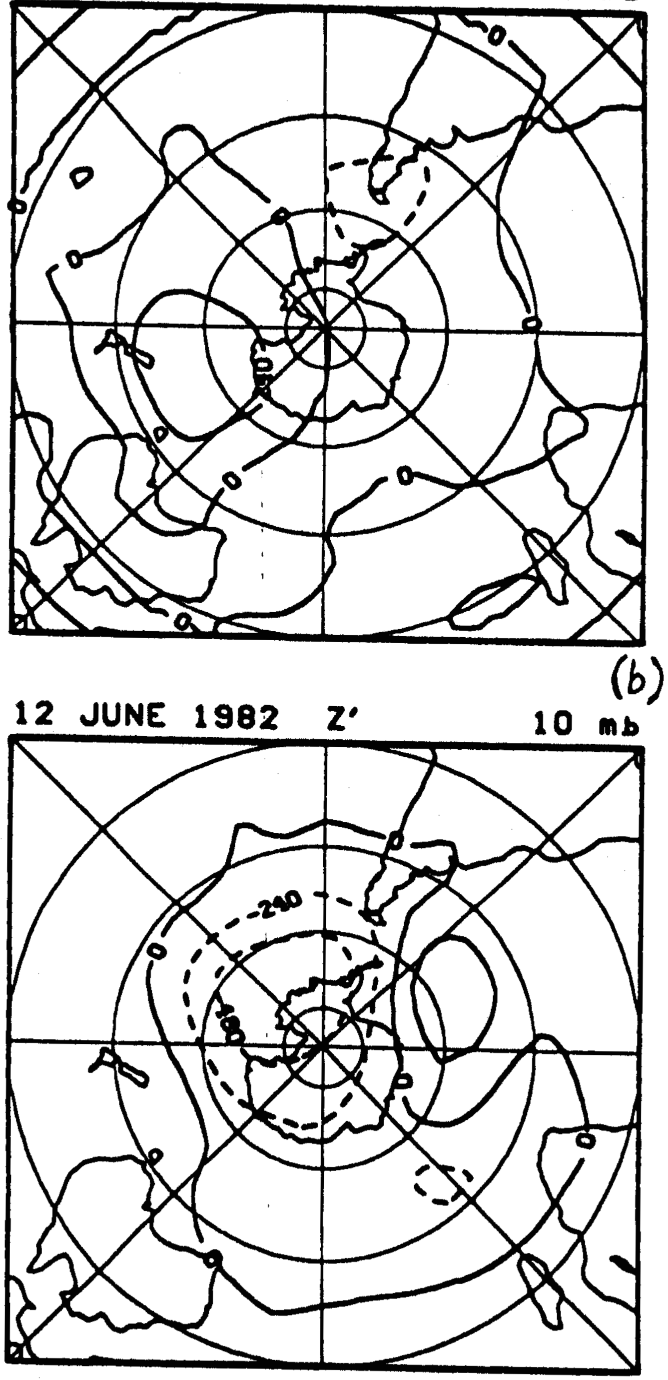



Fig. ~~13~~ 14

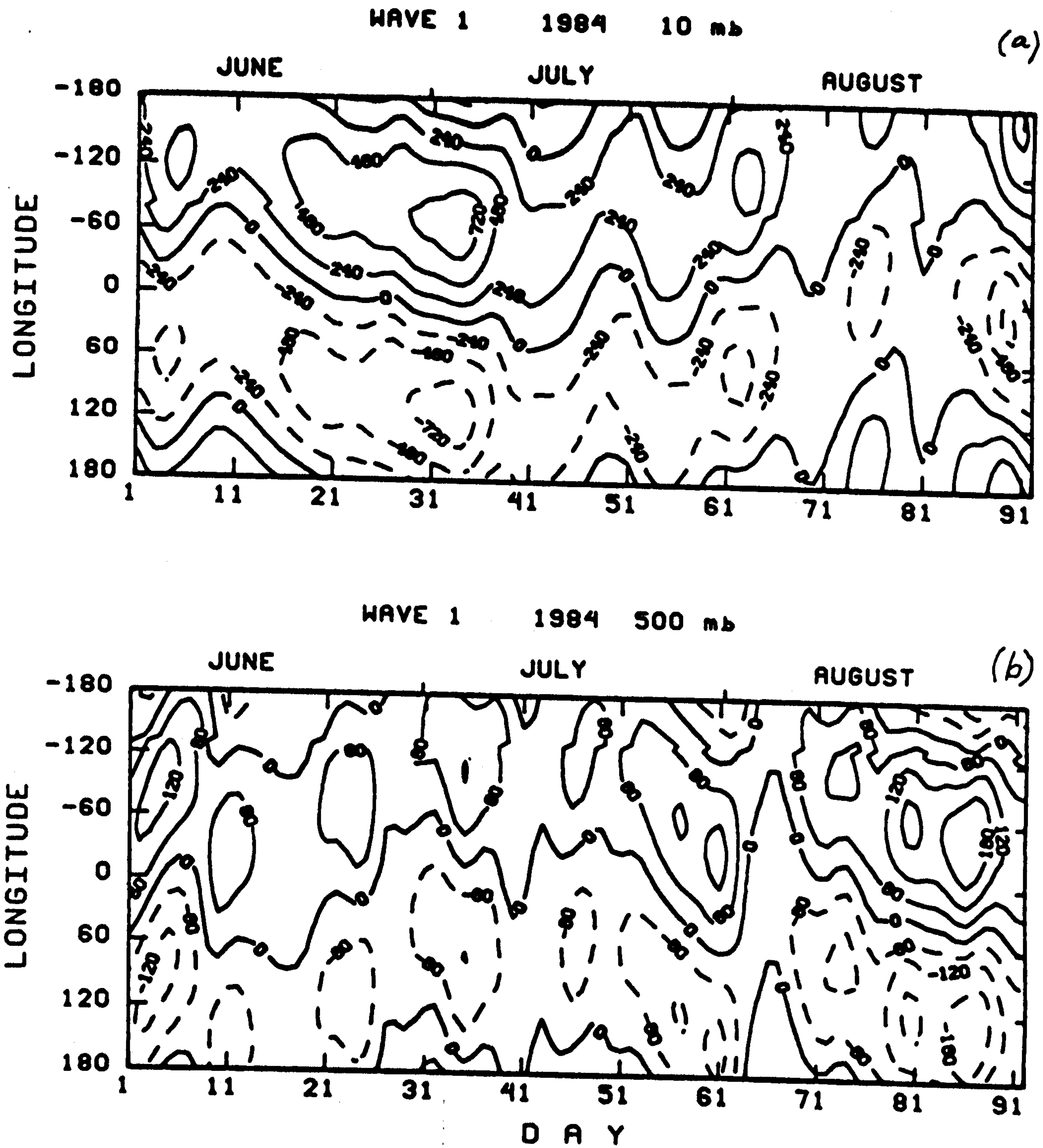

WAVE 1 1984 10 mb
(a)
JUNE
JULY
AUGUST
LONGITUDE
WAVE 1 1984 500 mb
(b)
DAY

Fig. ~~14~~ 15